PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B

Phil. Trans. R. Soc. B (2006) **361**, 1721–1731
doi:10.1098/rstb.2006.1900
Published online 7 September 2006

Physical conditions on the early Earth

Jonathan I. Lunine*,†

Istituto di Fisica dello Spazio Interplanetario, INAF ARTOV, Via del Fosso del Cavaliere, 100, 00133 Rome, Italy

The formation of the Earth as a planet was a large stochastic process in which the rapid assembly of asteroidal-to-Mars-sized bodies was followed by a more extended period of growth through collisions of these objects, facilitated by the gravitational perturbations associated with Jupiter. The Earth's inventory of water and organic molecules may have come from diverse sources, not more than 10% roughly from comets, the rest from asteroidal precursors to chondritic bodies and possibly objects near Earth's orbit for which no representative class of meteorites exists today in laboratory collections. The final assembly of the Earth included a catastrophic impact with a Mars-sized body, ejecting mantle and crustal material to form the Moon, and also devolatilizing part of the Earth. A magma ocean and steam atmosphere (possibly with silica vapour) existed briefly in this period, but terrestrial surface waters were below the critical point within 100 million years after Earth's formation, and liquid water existed continuously on the surface within a few hundred million years. Organic material delivered by comets and asteroids would have survived, in part, this violent early period, but frequent impacts of remaining debris probably prevented the continuous habitability of the Earth for one to several hundred million years. Planetary analogues to or records of this early time when life began include Io (heat flow), Titan (organic chemistry) and Venus (remnant early granites).

Keywords: Hadean Earth; Earth, origin; impacts; planetology; Archean Earth

1. INTRODUCTION

The earliest history of the Earth—commonly called the Hadean period—is barely recorded in the geologic record, and yet it is the time when liquid water appeared on our planet, and might—but not necessarily—have been the time when life began. The difficulties associated with determining the Hadean record are likely to remain regardless of the extent of field investigations, because the conditions postulated to be present during that time preclude significant preservation of the geologic record. On the other hand, the study of the Moon, planetary analogues to the early Earth and theoretical investigations all provide insights into the processes operating at this time. Understanding the environment within which the Earth formed also provides an important perspective on the nature of conditions on the earliest Earth.

The present review aims only to sketch the key issues associated with understanding conditions on the earliest Earth, beginning with the origin of the Earth and continuing with the sources of Earth's water, the effects of the collisional origin of the Moon and the prodigious Hadean impact rate. It continues with constraints and speculations from the literature on conditions present on the Earth's surface during the Hadean, and briefly outlines early Archean plate tectonics and evidence for granitic platforms during that time. Finally, the review identifies planetary bodies that provide potential analogues for environments and processes on the early Earth, and the implications for the environments within and mechanisms of the origin of life.

2. FORMATION OF THE EARTH

The formation of the Earth is thought to have been a multi-step process requiring tens of millions of years. The physical reasoning behind this number is straightforward: within the protoplanetary disk of gas and dust, as bodies grew in size, the mean spacing between them increased. Accretion (growth by sticking of particles) initially was rapid, with collisions frequent and relative velocities slow. By the time objects reached size scales of tens to hundreds of kilometres, growth to the next order of magnitude took millions of years, and when bodies reached a size between that of the Moon and Mars, they were effectively isolated from each other and growth stopped. Mutual perturbations among these widely spaced bodies would have eventually reshaped the circular orbits of these objects allowing them to collide, but only after hundreds of millions to billions of years (Levison & Agnor 2003). This time-scale is too long, because the time-scale for formation of the Earth's core—generally assumed to represent the accretion of a significant fraction of the Earth's mass (but see Halliday 2004*a*), is between 10 and 30 million years (Cameron 2002).

The resolution of this paradox is generally ascribed to the early presence of Jupiter, whose large gaseous envelope (270 out of 310 Earth masses total in the planet) must have been gravitationally captured within 1–10 million year time frame during which abundant gas is present around the forming Sun (Najita & Williams 2005; Pascucci *et al.* in press). The presence

*jlunine@lpl.arizona.edu
†On leave from Lunar and Planetary Laboratory, University of Arizona, Tucson, AZ 85721, USA.

One contribution of 19 to a Discussion Meeting Issue 'Conditions for the emergence of life on the early Earth'.

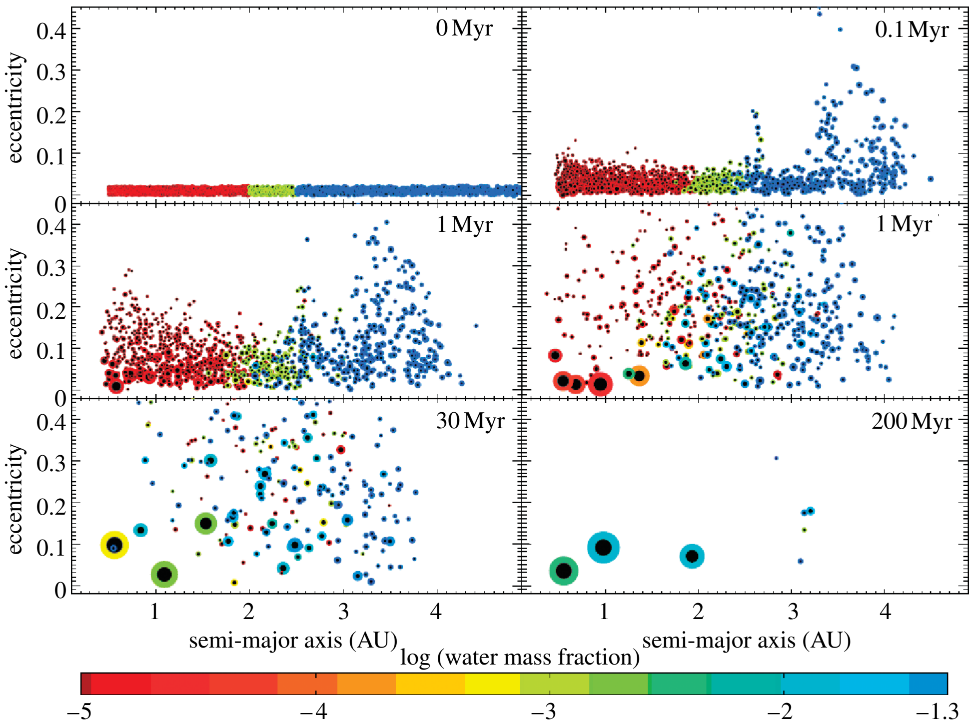

Figure 1. Example of the formation of terrestrial planets from a series of asteroidal-to-Mars-sized bodies using a symplector integrator simulation as described in the text. Six snapshots in time show the eccentricity versus semi-major axis of 1885 objects that collide, coalesce and grow under the perturbing influence of Jupiter and of their own growing gravitational fields. The size of each body is proportional to its linear diameter as it grows, and the amount of iron it contains is shown in black. Colours allow the eye to track mixing across regions of the solar system, and can also be considered a rough indication of the amount of water assumed present in these objects at the beginning, and in the bodies as they collide and grow, with the water mass fraction scale shown at the bottom. The final system is not precisely our own terrestrial planet system, but similar, and outcomes vary dramatically as boundary conditions are changed. From Raymond *et al.* (2006).

of Jupiter at or near its present orbital radius of 5.2 astronomical unit (AU) perturbs the orbits of the rocky planetary 'embryos' (a somewhat awkward term now embedded in the planetary formation literature) so that they intersect and collide with each other. This reduces the time-scale for the final stage of growth, from lunar- and Mars-sized to that of the Earth, to a few tens of millions of years (Chambers 2001).

Simulation of the formation of the Earth from smaller objects still possessing significant gravitational attraction requires specialized computer codes that are capable of correctly solving Hamilton's equations of motion taking into consideration the diverse time-scales involved, and the subtle gravitational resonances that develop between Jupiter and the growing planetary embryos. The codes, called 'symplectic integrators', are now in widespread use (Chambers 2003). In general, they are particularly useful in characterizing the sensitivity of terrestrial planet architectures (planetary mass, orbital semi-major axis and eccentricity) to the assumed initial conditions of embryo mass versus semi-major axis, and presence (and orbital distribution) of asteroidal-sized material. They are also useful in characterizing the radial mixing of material in the protoplanetary disk, as embryos from a range of semi-major axes collide with each other. The mixing alters the final composition of the Earth relative to what would have been obtained from material that originated at or near 1 AU from the forming Sun (figure 1).

However, the stochastic nature of planet formation makes it impossible to identify a particular event (a collision between two embryos) with specific known episodes during formation of the Earth, or even with a final particular planetary architecture in which one planet might be identified with Earth, another with Venus, etc. For example, the giant collision that formed the Moon is 'reproduced' in the simulations in the sense that collisions between embryos might involve bodies in closely similar orbits (a prerequisite for getting the right lunar geochemistry; see Halliday 2004*b*). However, one can neither identify a particular collision with the actual Moon-forming one, nor track the detailed parameters of the collision, with symplectic integrators; other codes that create and track a specific collision based on a prespecified set of initial conditions are required (Canup 2004).

The symplectic simulation of terrestrial planet formation represents a significant advance in our understanding of how terrestrial (rocky) planet systems form, and Earth-sized planets grow, relative to earlier codes that failed to conserve the Hamiltonian function and correctly account for subtle resonant interactions. However, there are limitations in these codes. In particular, the precise architecture of our terrestrial planet system is not reproduced in detail. In general, it is difficult to finish with two Earth-mass planets (Earth and Venus) and two planetary embryos (Mercury and Mars). More importantly, the low eccentricities of the Earth and Venus are not obtained in simulations that do not include a large population of smaller objects along with the planet-sized embryos. However, the symplectic integrations cannot explicitly include all the numerous small bodies that semi-analytic models predict were present (Goldreich *et al*. 2004), owing to limitations of computing power. The final eccentricities of bodies close to Earth's mass and semi-major axis remain slightly above the observed values, probably for this reason (O'Brien *et al*. in press).

In addition, at issue is whether collisions between planetary embryos result in net accretion or disruption of the colliding bodies. The assumption to date has been that the larger of the two colliding bodies generally grows, but under some conditions a significant amount of material may be ejected (Asphaug *et al*. 2006). This has two potential consequences. First, the total accretion time may be extended, although to what extent is not yet known. Second, a large population of smaller bodies should be generated by the mega-collisions, and these may contribute significantly to the reduction of the final eccentricities of the terrestrial planets. Ultimately, advanced codes might track the evolution of the smaller debris—the raw material for early impacts on the Earth discussed below—but at present this is not possible.

The principal utility of these models is twofold. First, the models emphasize the importance of so-called giant collisions between embryos as the dominant mode of growth of the terrestrial planets. Such impacts deposit heat deep inside the growing planets, and therefore maximize the overall temperature rise experienced by the Earth during formation, relative to accretion of smaller bodies. Second, such models allow evaluation from a dynamical standpoint of various hypotheses for the source of Earth's water and carbon, which are explored in the following section.

3. SOURCES OF WATER AND ORGANICS FOR THE EARTH

The Earth today contains between several and ten times the amount of water contained in the oceans, with the upper limit extremely uncertain (Abe *et al*. 2000). The source of Earth's water remains unknown, the possibilities including: (i) objects local to the 1 AU region, (ii) bodies formed further from the Sun and hence colder than those at 1 AU and (iii) direct transfer from the nebular gas. Based on astronomical observations, the third of these possibilities can be ruled out by the mismatch between the lifetime of nebular gas and the time-scale for terrestrial planet formation constrained by radioisotopic systems (Cameron 2002; Halliday 2004*a*). At the time the nebula dissipated, the Earth would have been too small to retain a sufficient amount of nebular water against giant impacts in order to explain the inventory of water.

A local source of water, option (i), is the most direct and simplest possibility. It requires that the nebula at 1 AU had been cold enough to directly condense water ice, bind it in rock as water of hydration or adsorb it onto mineral grains at parts per thousand (by number) levels. Hydration of silicates may have resulted from reaction with ice, either directly condensed (if the snowline was inward of a particular object), acquired through collision with icy bodies on eccentric orbits or spiralling inward via gas drag (Cyr *et al*. 1998), or directly through reaction of water with the nebular gas (Ganguly & Bose 1995).

Traditional nebular models predict temperatures too warm at 1 AU for condensation or hydration throughout the time gas was present (Cassen 2001), although it is conceivable that adsorption occurred (Stimpfl *et al*. 2004). However, observations of denser, colder disks in which planetary systems may be forming hint that the minimum point of water ice condensation (the so-called 'snowline') in the nebula could have been inward of 2 AU, perhaps as close as 1 AU (Sasselov & Lecar 2000). At 1 AU, there would have been bound water of hydration in silicate minerals, equivalent to the several per cent (by number) measured in carbonaceous chondrites-meteorites almost universally regarded as having come from parent bodies beyond 2.5 AU and showing aqueous alteration indicative of liquid water within these parent bodies.

The principal difficulty with this source is that the chondrite meteorite classes are matched to asteroidal parent bodies whose radial distance of origin from the Sun does not correspond to the colder, more inward snowline inferred from the dense cold disks. The meteorite classes, when assigned to asteroidal parent bodies, imply a declining water of hydration with decreasing solar distance that, extrapolated to the Earth's orbit at 1 AU, imply that the material from which the Earth formed was quite dry. The extreme dryness of the proto-Earth material is consistent with the extreme volatile-poor state of the Moon and its inferred collisional parent body, which may not be entirely explainable through the impact process itself (Halliday 2004*b*). Frustratingly, there is no sample in the known collection of meteorites that arguably represents a primitive sample of the Earth, i.e. an 'Earth chondrite', which would resolve the issue. The enstatite chondrites are very dry and have just enough water that, were there no loss of water owing to giant collisions, could explain the Earth's crustal water (and barely its mantle water), but simulations of the collisional process argue that significant amounts of water are indeed lost in the giant impacts (Genda & Abe 2005). It would appear that an outward source of water is required unless an unexpected upswing in water of hydration was characteristic of the missing Earth-chondrite class. It is difficult to invoke planetesimals at 1 AU being wetter than the enstatites in the face of the evident gradient among chondrite classes.

A distant, outward source of water, derived from material at least beyond 1.5 AU, required radial mixing among planetary embryos and probably the population of smaller mass bodies as well. Comets are by far the most numerous and possibly, in the aggregate, most massive source of water for the terrestrial planets, but numerical simulations show that it is very difficult to scatter such material inward so that it collided with the Earth; most of the icy bodies near and beyond (greater than 4 AU) the orbit of Jupiter were quickly scattered outward to the Oort Cloud and, to a lesser extent, the Kuiper Belt (Morbidelli *et al*. 2000). However, they may have been a key source along with numerous asteroidal-sized bodies of Martian water (Lunine *et al*. 2003).

The deuterium-to-hydrogen ratio (D/H) in terrestrial ocean water, 150 parts per million roughly, is slightly enriched over the value in the Earth's mantle but is thought to be roughly indicative of the value in the primordial sources of water. The value is half that seen in three short-period comets, and therefore supports the proposition that comets were *not* the main contributor to the Earth's total water budget. Carbonaceous chondrites have deuterium in bound water that ranges over broad values, but centres near the oceanic value (Robert 2001). The parent bodies of the carbonaceous chondrites are asteroids beyond 2.5 AU, the remnant of a much more massive population of planetary embryos that once occupied that region.

At issue, then, is whether the dynamical models described in §2 predict that a sufficient number of planetary embryos from beyond 2.5 AU can reach the Earth to supply its required inventory of water—at least five or so times the mass of the present ocean, and probably double that number in order to account for impact losses. In fact, a significant fraction of the simulations predict that enough carbonaceous or ordinary chondritic planetary embryos—the material assumed to dominate between 2 and 4 AU—reaches the Earth to provide at least the lower end of the required water abundance, while at the same time the vast majority of material that formed the Earth is still accreted from 1 AU (Morbidelli *et al.* 2000). However, there are some caveats. First, it is difficult to arrive at very wet mantles, i.e. tens of Earth masses of water, as some geochemists have argued the Earth once possessed (Abe *et al.* 2000), particularly in view of significant impact loss amplified by the very presence of surface liquid oceans (Genda & Abe 2005). Second, carbonaceous chondrite parent bodies have oxygen isotopic and siderophile abundance patterns distinct from that of the bulk Earth, and therefore not more than 1–3% of the Earth could be composed of material from such bodies, depending upon (respectively), whether the chondritic material was added after or before core formation (Drake & Righter 2002). The lower limit is certainly too low to account for the Earth's inventory of water, and the upper value may be only marginally generous enough provided the Earth sits near the dry end of its possible range of water values, and loss by impacts is minimal. In view of the fact that the upper value pertains to accretion prior to core formation, when the Earth was perhaps significantly less massive than today, loss of material (including water) during collisions with other embryos may have been too extensive (Asphaug *et al.* 2006) to admit the carbonaceous chondritic source.

There may be ways out of this dilemma while retaining the hypothesis that the material supplying water was carbonaceous chondritic. The oxygen isotopic constraint, which considers the correspondence between that for the Earth and the Moon, assumes that there was no late rehomogenization of the value between the two bodies during the collision that created the Moon. This assumption was probably violated based on simulations that demonstrated very high temperatures reached by the material during collision (see the colour plates detailing such collisions in Canup 2004). The siderophile constraint is valid only if the embryos that delivered water to the Earth were not differentiated. If they were differentiated, then most of the siderophile elements were confined to the cores and these were not effectively mixed by the collisions into the Earth (Cameron 2002). Since most of the mass delivered to the Earth was in the form of lunar-to-Mars-sized bodies, these were almost certainly at least partly differentiated and hence the alteration of the Earth's siderophilic abundances was far less efficient than previously assumed (Drake & Righter 2002). Another alternative is that the population of hydrated bodies beyond 2.5 AU included carbonaceous chondritic bodies but the bulk of the material there had a different set of isotopic and elemental abundances. This is *ad hoc* but cannot be formally ruled out because the vast majority of material once in the asteroid belt is gone. The ordinary chondrite source, more compatible with terrestrial composition, is yet another possibility that new simulations (O'Brien *et al.* in press) are admitting, and work is in progress to evaluate the plausibility of this source.

The origin of terrestrial organic molecules is less clear because the isotopic constraints are less secure. Again, carbonaceous chondrites are rich in organic molecules, and their delivery to the Earth as large planetary embryos from beyond 2.5 AU would have provided the Earth with abundant enough carbon to account for its roughly 10^{-4} abundance by number in the mantle and the deep crust. But late addition of comets—while only a minor contributor of water—could have been an important source of surficial and shallow-crustal carbon (Chyba *et al.* 1990), added after the last giant collision so that it would not have been ejected subsequent to deposition. A weak indication that the addition of Earth's observable carbon was from a steady stream of smaller material, rather than the stochastic strike of a few larger bodies, comes from the rough (factor of three) correspondence in amount of the carbon dioxide in the Venusian atmosphere—derived in ancient times from that planet's crustal carbon (Kasting 1988)—with the total amount of shallow crustal carbon on the Earth.

4. FORMATION OF THE MOON

While the dynamical simulations of terrestrial planet formation indicate that growth from lunar- and Mars-sized bodies involves a series of giant impacts, the effect of such impacts on the growing Earth was highly varied depending on the mass of the impactor, the mass of the proto-Earth at the time of collision, the relative velocities between the two bodies, the angle of incidence of the impactor (Canup & Pierazzo 2006), the state of the Earth's surface (water-covered versus dry; molten versus solid) and the composition of the impactor (e.g. wet or dry). The origin of the Earth's Moon lies in one such giant collision, and one with very specific properties initially constrained by geochemistry (Hartmann & Davis 1975), and later by detailed dynamical simulations (Benz *et al.* 1986; Canup 2004).

The Moon-forming impact occurred towards the middle to late time in the growth of the Earth, principally based not only on isotopic data (Halliday *et al.* 2004*a*), but (more weakly) also on the need for

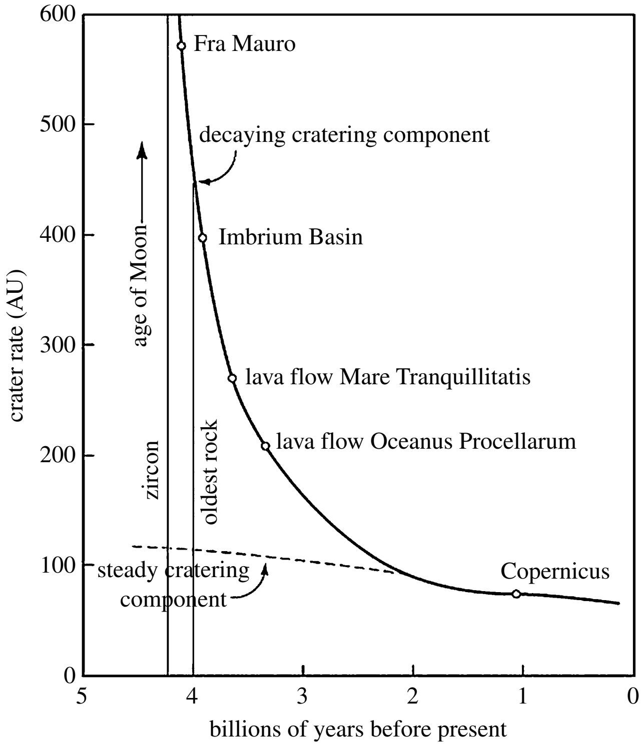

Figure 2. Schematic change in cratering rate over time in the Earth–Moon system, with major events based on dating of Apollo samples labelled. Vertical lines indicate the ages of the oldest terrestrial zircon, and whole rock, samples. Modified from a variety of sources and originally published in Lunine (2005).

the Earth to have been well-differentiated at the time of the collision. The event involved a colliding body the mass of Mars or a few tens of per cent larger, coming in obliquely so as to produce material that would collide with the Earth a second time so that significant material would remain in orbit around the Earth rather than ejected into solar orbit (Canup 2004). The geochemistry of the Moon implies that the impactor was close in composition to that of the Earth, hence probably had an orbit close to that of the Earth for much of its own growth, and this is consistent with the need for an oblique rather than direct collision.

If the Moon-forming impact was late in the history of the Earth's growth, a hydrosphere including a surface liquid-water ocean may well have been present. Much of it would have been vaporized and lost during the collision (Genda & Abe 2005) along with surface organic molecules whether pre-biotic or alive. Thus, the formation of the Moon was probably a late, if not the last, seminal event in which the Earth's surface was strongly heated, dehydrated and sterilized. Restoration of the hydrosphere would presumably have required further outgassing, since there is weak evidence that the impactor itself was dry, and even if not much of the retained water should have been deeply deposited within the Earth.

Once established in orbit, the Moon would have provided a strong tidal perturbation on the Earth, since its orbit was initially much closer to the Earth than it is today, and the obliquity of the Earth's spin—if not stabilized by the rapid rotation rate—would have been locked and stabilized at or near the present value (Ward 1973). Whether either of these effects was important in the subsequent origin of life is unclear, because we know almost nothing about the details of the terrestrial environment in which life formed. In contrast, the subsequent effects on life's evolution were strong and have been well-documented in many places, including the popular literature (Ward & Brownlee 2000).

5. THE POST-FORMATION IMPACT ENVIRONMENT

After the final giant impact of lunar-to-Mars sized bodies into the Earth, our planet continued to be pelted with debris up to the present. Studies of lunar impact craters and returned lunar samples gives an impact rate for the Earth–Moon system shown schematically in figure 2. Impacts of a size sufficient to produce the large impact basins seen on the Moon and Mars were probably the limiting factors in when life gained a permanent toehold on the Earth (Sleep *et al.* 1989). While the largest impacts—involving bodies several hundreds of kilometres across—may have been sterilizing by boiling most or all of the surface water (whether crustal or exposed as an ocean), they were relatively few in number. The lack of preserved impact basins on the Earth prevents the specification of the time of the last such impact—in contrast to a fairly well-specified time of the lunar-forming giant impact.

Smaller impactors—of order 100 km or less in size—while destroying the organic inventory of particular local areas, might have created post-impact environments that were particularly favourable for the origin of life, by creating large surface areas for reaction, introducing important trace minerals and synthesizing mineral surfaces appropriate for template-driven chemistry (Cockell 2006). Furthermore, oblique impacts of smaller (1–10 km scale) impactors leave a relatively lightly shocked or unshocked region in the wake of the impact plume, preserving or only lightly altering organic deposits (Artemieva & Lunine 2005). Thus, except for the largest basin-forming impacts (of which Earth might have zero or a small number), impacts might not have impeded or even helped organic synthesis processes involved in life's origin.

The survival of living organisms in impacts is suggested by studies of bacteria collected after impacts of sounding rockets at velocities of roughly 1 km s^{-1}, in which up to several per cent survival rate is seen (Fajardo-Cavazos *et al.* 2005). Most impacts on the Earth had larger impact velocities—by an order of magnitude and hence two orders of magnitude in deposited energy—but a very small fraction of organisms surviving modestly sized impacts cannot be ruled out. The importance of these issues lies in the evidence (discussed in §6) that water was stable on the Earth at 4.0 billion years or earlier, a time when the entire surface of the Earth experienced repeated impacts of moderate size. Was the Earth habitable at that time? Had there been enough time for life to form and become robust enough to survive environmental excursions associated with moderate-sized impacts? What is the relationship between the macroscopic effects of impact processes on global and regional scales, and microenvironments at varying distances from the site of the impact? No real evidence exists to guide modelling of sufficient fidelity to confidently address these questions.

At best, one can say that the question of impact sterilization of life remains a difficult one, and certainly the last major impact basin-forming event on the Moon at 3.8 billion years is not a hard limit—in either direction—for the age at which the Earth finally became continuously habitable.

6. CONDITIONS IN THE HADEAN WITH RESPECT TO THE ORIGIN OF LIFE

The key pieces of evidence for conditions in the Hadean come from a very few samples of crust that have survived to the present day. Chemical remains in the form of 'zircons' suggest liquid water in the upper crust of the Earth, if not the surface, as early as 4.4 billion years ago. Zircon is short for zirconium silicate ($ZrSiO_4$), a mineral that is found primarily, but not exclusively, in volcanic or intrusive rocks more felsic than basalts. Zircons are of interest in Earth history because they can be readily dated using the uranium–lead (U–Pb) radioisotopic system, they are resistant to melting during metamorphism in the crust, and robust against mechanical destruction during sediment transport. Furthermore, oxygen is tightly bound in zircons, and this lack of mobility means that the oxygen isotope ratio in these minerals is hardly altered over time. For these reasons, a group of ancient zircons found in a complex of hills on a very ancient and stable piece of Western Australia, which have U–Pb dates between 3.9 and 4.4 billion years before present, are of keen interest (Mojzsis *et al.* 2001).

The rocks bearing the zircons consisted of gneisses, rocks that have been heavily metamorphosed, with compositions ranging from that of granites to tonalites. The zircons that exceed 4 billion years in age are found in a conglomerate of quartz pebbles, and were not formed there. None of the rocks within which the zircons were found seem to be the original rocks within which the zircons were formed; the zircons outlasted their host rocks and were deposited (perhaps repeatedly) in younger sediments, which later were metamorphosed.

The ratio of the stable oxygen isotopes ^{18}O and ^{16}O in the zircons provides an indication of the mineral's formation environment. This ratio is well determined for the mantle of the Earth, and eruption of basalts from the mantle, with varying ages, show little change in the oxygen isotope ratio (Halliday 2004*a*). On the other hand, crustal rocks show a wide variation in the oxygen isotope ratio, and in particular crustal material that is repeatedly recycled in the presence of water to form rocks of granitic or similar composition have significantly higher values of the oxygen isotope ratio.

Measurements of oxygen isotope ratios in younger zircons, still embedded within the rock from which they formed, yields a well-known relationship between the oxygen isotope value in the zircons and that of the host rock. In the case of the ancient Australian zircons, the measured oxygen isotopic ratios imply host rock values significantly higher than for mantle rock, and consistent with crustal rocks that had interacted with liquid water (Mojzsis *et al.* 2001). The liquid water need not have been at the surface, and indeed more likely was beneath the surface in hydrothermal systems where intimate and repeated contact with minerals in the crust raised the oxygen isotopic ratios of the latter, including that of the zircons. How deep the liquid water might have been cannot be determined, but was definitely within the crust rather than deep within the mantle, where a different oxygen isotope ratio would have been obtained.

About 1% of the zircons in the sample were dated to 4.3 billion years before present, with one fragment (Jack Hills) at 4.4 billion years (Peck *et al.* 2001) and they imply that liquid water was at least circulating within the crust, if not at the surface, at that very early time. The presence and composition of the zircons also implies a surrounding rock from which it formed that was not as mafic as basalt, but rather more continental in nature, such as might have differentiated from the mantle in a tectonic setting equivalent to present-day Iceland (Valley *et al.* 2002), in which large basaltic islands (submerged or exposed) enclosed small granitic masses (Halliday 2004*a*).

However, the inference that continental-type crust therefore existed at 4.4 billion years is controversial, since this would imply a tectonic setting that must be regarded as surprisingly evolved geochemically for the early Hadean. Alternatively, it is possible that the zircons formed in a high-silica environment, but do not necessarily represent continental or even Icelandic-type crust (Moorbath 2005). In any event, very thick continental crust like that of today would have been difficult to sustain early in Earth's history in view of its concentration of uranium, thorium and potassium, for the high (4–5 times present) Hadean values of their radioactive isotopes should have led to melting in the granites and rereaction with other rocks (S. Moorbath 2005, personal communication). The oldest whole rock samples, sedimentary or metasedimentary, are dated to 4.0 billion years ago, and from that point

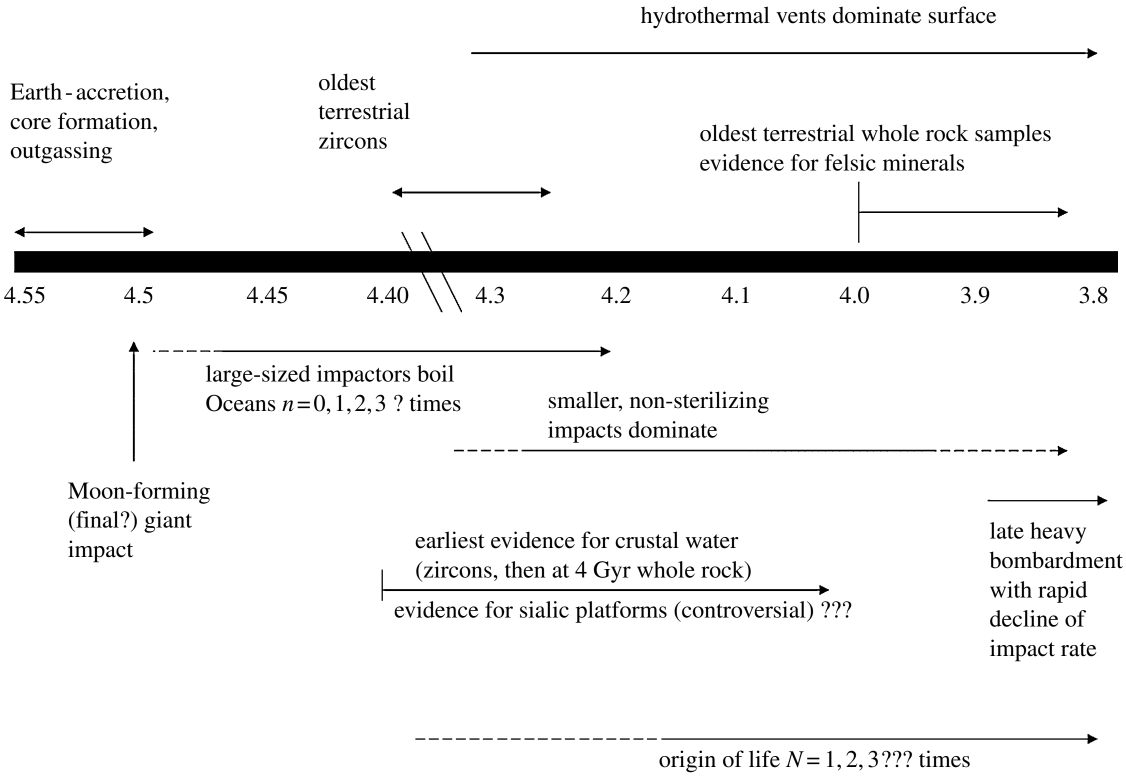

Figure 3. Speculative integrated early history of the Earth covering the first 700 million years from the formation to the tail-off of the impactor flux. During this time, the Earth and Moon form, the Earth acquires its water and carbon content, oceans and hydrothermal vent systems appear, and life begins. The hydrosphere-destroying megaimpacts give way to sterilizing impacts and then, at the end of the timeline, non-globally lethal impacts that might have created habitable hydrothermal systems.

onward increasing evidence for felsic minerals exists, such as in the Acasta gneiss complex (Bowring & Williams 1999).

A different record of oxygen isotopes, those in cherts, has the potential to constrain surface temperature back to 3.5 billion years ago. Cherts are a sedimentary form of silica (SiO_2) in either very small crystals (fine-grained) or amorphous (glassy) form, precipitating directly out of rivers or ocean waters, or forming from rocks that are subjected to mild increases in temperature and pressure. Biogenic chert is the most commonly found form, in which radiolarians and diatoms secrete silica, and then lithify in the form of very fine grained, microcrystalline or glassy quartz.

The key to cherts as an indicator of temperature is that the oxygen isotopic content of the chert bears a definite relationship to that of the environment in which it is precipitated. If precipitation occurs in an ocean environment, then the ratio $^{18}O/^{16}O$ of the silica decreases with increasing temperature in a manner that has been determined in the laboratory. Hence, cherts act to record the ambient water temperature through the oxygen isotopic enhancement during its formation. Further, somewhat akin to zircons, cherts tend to be well preserved in the sedimentary record over billions of years.

Because cherts form in so many different kinds of sedimentary environments, each of which may generate a different relationship between the oxygen isotopic ratio and the ambient conditions, it is not straightforward to use cherts as a 'geothermometer'. In some environments, the oxygen isotope ratio may be altered in ways that have nothing to do with the surface temperature. The argument that cherts reflect the oceanic temperature in some way is that processes during or after formation will tend to lower the $^{18}O/^{16}O$ value in cherts relative to the value obtained during precipitation of the silica from the water. Therefore, for a collection of cherts of a given age, the cherts with the highest $^{18}O/^{16}O$ values should most nearly reflect equilibration with ocean waters during formation, and have suffered the least alteration. Knauth (1992) has derived an Archean oceanic temperature of 50–70°C on this basis. Whether this implies the Hadean was even hotter is highly uncertain, since the low solar luminosity would have made global ocean and atmospheric temperatures extremely sensitive to the abundances of various greenhouse gases (Pavlov *et al.* 2000).

An alternative explanation for the derived temperature is that it reflects alteration of the oxygen isotopic ratio of the cherts in water in hydrothermal systems at the base of the Archean ocean, and would then point to the ubiquity of such systems present at the time (Kasting 2006). The chert data then support the notion that there were hot crustal environments in the Archean, and the higher geothermal heat flow of the Hadean would argue for even more at that time. However, in this case, the climate away from such hydrothermal systems need not have been especially hot, as has been argued on the basis of other geochemical indicators (e.g. Sleep & Hessler 2006), or even cold, based on the problem of providing sufficient greenhouse gases to counter the faint early Sun (Sleep & Zahnle 2001).

Figure 3 offers a timeline of events on the early Earth consistent with the discussion presented above. Left open are many questions, including when the last planetary-sterilizing impact occurred, and when permanent exposures of continental (or basaltic-island) structures stood above the ocean, as opposed to remaining submerged. The geochemistry of crustal materials exposed to hydrothermal and surface aqueous processes was different from that of the present day, but the specifics of these differences remain debated. Further fieldwork to discover mineral and whole rock remnants of the Hadean (a fruitful yet paradoxical effort given the original definition of the Hadean as the time 'before' the rock record) will provide a perspective more detailed and more secure than can be provided by theory. Most exciting would be the discovery of faint chemical traces of biological processes in rocks at or prior to 4.0 billion years ago. While the search for the Hadean rock record continues to bear fruit, an additional perspective is provided by the wide range of planetary environments elsewhere in the Solar System that are informative with respect to analogous processes, and possibly with respect to preserving direct geologic clues not available on Earth.

7. PLANETARY ANALOGUES AND CLUES TO THE EARLY EARTH ENVIRONMENT

(a) *Venus*

Earth's nearest neighbour is the planetary body that most closely resembles Earth in terms of mass, size and—as far as is known—composition. The very large deuterium-to-hydrogen ratio in the present Venusian atmosphere strongly implies that the planet lost oceans worth of water, (Donahue 1989) through a solar-driven evaporation and photolysis process (Kasting 1988), probably (but not necessarily) early on. After water loss (and possibly before) the expression of interior heat flow was not through terrestrial-type plate tectonics but rather vertical tectonics involving fixed basaltic volcanism and flaking of the base of the crust into the mantle (Phillips & Hansen 1994). However, if water was present long-enough early in its history, Archean or Hadean-style plate tectonics might have occurred, and finding evidence for ancient plate tectonics is a key but difficult goal of Venus exploration. The global episode(s) of basaltic volcanism have covered most of the surface with thick flows, but it is conceivable that ancient crust preserving the earliest tectonic history is exposed somewhere, perhaps in a highland area. It would be extraordinarily important to find granitic rocks on Venus; their age dating might help pin down how soon after the crust formed that non-basaltic, more felsic rock could have appeared on the Earth-sized planet.

Detailed mineralogical mapping below the obscuring clouds and strong carbon dioxide absorption bands is essential for deciding where to go sample on Venus. Both the global reconnaissance and the surface sampling will be extremely challenging, requiring operation in high temperature (400°C) and high pressure (90 atmospheres) environments.

(b) *Ceres*

If the Earth's water was acquired in part or in whole from the asteroid belt, then exploration of the asteroids directly is of keen importance in understanding the sources of Earth's water. The largest asteroid, Ceres, orbits at 2.8 times the Earth–Sun distance, and exhibits spectral features that suggest the presence of hydrated minerals, and hence of water (McCord & Sotin 2005 and references therein). However, little else is definitively known about this object, including its possibly low density, early heating history and total extent of water. Like other asteroids, direct exploration by spacecraft will be required to infer the details of its history, and this is planned for Ceres on the US–Italian–German Discovery mission Dawn, set for launch in 2007. However, the remote sensing investigations planned will likely not be enough to provide a definitive indication of water content, since Ceres is large enough to have undergone differentiation and outgassing. It is quite possible that eventual drilling or even impact–extraction of subsurface material will be required to probe for water.

(c) *Io*

Jupiter's innermost Galilean satellite Io, about the same mass as the Earth's Moon, is heated by the tidal forces exerted by Jupiter, and hence an overall heat flow of approximately 3 W m^{-2} (Veeder *et al.* 2004)—30 times the mean oceanic heat flow on the Earth at present and three times the peak heat flow at the mid-ocean ridges (Stein & Stein 1994). Thus Io's heat flow is comparable to or exceeds the terrestrial Hadean heat flow, and its intensive volcanism might provide insight into mechanisms of crustal heat flow at high values. The ability to use this object as a guide to terrestrial Hadean tectonics is limited by the virtually complete absence of water on and within Io, the smaller size of the body, and possibly as well by the different interior distribution of heat flow caused by the strong tidal heating versus Earth's radiogenic heating. However, close up or *in situ* study of Io is practically impossible owing to the enormously intense radiation environment in which it is embedded, caused by nearby Jupiter's powerful magnetic field.

(d) *Europa*

Next out from Io is Europa, a similarly sized rocky moon encased in a shell of water roughly 100–200 km thick. The Galileo spacecraft provided indirect evidence that beneath the icy crust is a liquid water ocean, leading to the possibility that habitable environments exist within. Extensive work has been done on possible hydrothermal systems at the interface between the liquid water and the rocky core beneath, exploring the differences with Earth determined by the potentially strong differences in mineralogy (Zolotov & Shock 2003). If Europa hosts life, the biomass is limited and metabolisms may be primitive (Chyba 2000). However, accessing the subsurface ocean will be difficult unless there are areas of the icy crust that are thin, and it will require an orbiter around Europa to determine crustal thickness through measurement of the body's tidal flexing (Wahr *et al.* in press). Again, the external radiation environment is both lethal to organisms and difficult for spacecraft operations, although not as severe as at Io. A possible alternative just recently discovered is that a subsurface pool of liquid water may exist under the surface of Saturn's much smaller moon, Enceladus, where a geyser of water and organic molecules was discovered by Cassini (Kargel 2006).

(e) *Titan*

Saturn's largest moon Titan is a world clothed in a dense nitrogen atmosphere with a surface temperature of 94 K and pressure of 1.5 bars. Methane is the second-most abundant gas with a mixing ratio of 5% near-surface, declining upwards as the temperature declines towards 70 K and clouds form above 8 km altitude (Niemann *et al.* 2005). Methane not only plays the role that water plays on Earth in terms of meteorology—but is also broken apart at very high altitudes by solar UV radiation; the hydrogen escapes and heavier hydrocarbons and nitriles are then formed from the radicals. These condense and descend to the surface as aerosols, residing then on the surface in the form of liquids (ethane, propane along with any original methane), and solids (other hydrocarbons and nitriles). The aerosols, once formed in the upper atmosphere, form a haze that, along with the deep methane absorption bands, has made examination of the Titan's surface difficult from Earth and spacecraft—use of near-infrared imagers with spectral discrimination along with radar is required. The joint US–European mission Cassini–Huygens began a complete investigation of this Mercury-sized world, made of rock and ice, in July 2004, with the orbiter making multiple flybys of Titan and an entry probe (Huygens) successfully descending to the surface and making measurements *in situ* for several hours in mid-January 2005. Titan's surface and atmospheric chemistry suggest a world that is subjected to a diverse range of geologic and atmospheric processes, including fluvial erosion, a possible type of volcanism involving water and an antifreeze agent such as ammonia, aeolian erosion and deposition and cloud/rain (methane) formation (Elachi *et al.* 2005; Tomasko *et al.* 2005). These processes seem to be ongoing, or recent, rather than ancient; in this sense Titan far more closely resembles the Earth than either Mars or Venus, even though the working materials and conditions on Titan are very different from those on the Earth.

Much has been made of Titan's possible role as an analogue to the early Earth before life, owing to its ongoing abiotic atmospheric (and possibly surface) organic chemistry, an analogue made imperfect by the virtual absence of atmospheric carbon dioxide and the far more reducing conditions therefore obtained in Titan's atmosphere and on its surface (Lorenz *et al.* 2001).

However, other analogies are also of interest. The origin of Titan's surface and atmospheric methane budget is thought to lie in outgassing of this volatile from the interior—possibly episodically (Tobie *et al.* 2006), since it is used up by the photochemistry and no large surface liquid reservoirs have been detected so far. The signature in Earth's atmosphere of its extensive early outgassing is the large amount of radiogenic argon, ^{40}Ar, one of the products of the radioactive decay of the potassium isotope ^{40}K. The Earth's inventory of ^{40}Ar in its atmosphere is about 50% of the maximum possible that could have been injected from the interior up to the present day, and is indicative of extensive outgassing. This outgassing included the crustal water we see today. The value of ^{40}Ar measured in Titan's atmosphere (Niemann *et al.* 2005) is equivalent to between a few per cent and 10% the maximum that Titan could possess, and suggests that Titan is moderately outgassed—not quite to the same extent as the Earth but of the same order. This is consistent with the notion that outgassing of volatiles such as methane has occurred through Titan's history, and might still be ongoing today.

The second analogy with the Earth has to do with hydrothermal environments generated by large impact craters, which may have produced localized habitable hydrothermal systems on the early Earth and Mars in locations where volcanic hydrothermal systems did not exist (Cockell 2006). The equivalent probably has existed on Titan as well. Two large impact craters have been identified in radar images (Elachi *et al.* 2005), one over 450 km across; this latter shows evidence for fluvial erosion and has a peculiar floor that suggests melting of the crust or infill with dark material. Calculations of the impact process (Artemieva & Lunine 2005) and refreezing of the water ice crust of Titan (O'Brien *et al.* 2005) indicated that liquid water would be generated in significant quantities and be stable—depending on possible antifreeze agents present—for thousands of years or more in such a large crater. Mixing of liquid water with ambient organics in such an impact-generated 'methanothermal' system might have generated amino acids and other compounds of interest to the origin of life. Sampling of interesting surface organics in future missions might well be most fruitful in and around this crater, or others that have yet to be discovered.

The landing site of the Huygens probe exhibited drainage channels probably carved by heavy methane rains and other channels cut into frature systems (Tomasko *et al.* 2005), as well as methane and ethane residing immediately beneath the surface at the landing site—detected by evaporation into the Huygens Probe's mass spectrometer (Niemann *et al.* 2005). Titan evidently is an active world in a way that we regard the Earth as active, and it will be interesting to explore it in the future to further investigate the analogous processes discussed above, and future analogies to the present and past Earth yet to be realized.

I am deeply grateful to Dr Angioletta Coradini and the Institute for the Physics of Interplanetary Space (Rome), which she directs, for hosting my academic sabbatical, during which the present article was written. I thank Steven Moorbath for his valuable comments and criticisms, and an anonymous referee for suggestions that led to an improved manuscript. My research on the origin of the Earth's volatiles is supported by the NASA Planetary Atmospheres Program.

REFERENCES

Abe, Y., Ohtani, E., Okuchi, T., Righter, K. & Drake, M. 2000 Water in the Early Earth. In *Origin of the Earth and Moon* (ed. R. M. Canup & K. Righter), pp. 413–433. Tucson, AZ: University of Arizona Press.

Artemieva, N. & Lunine, J. I. 2005 Impact cratering on Titan II. Global melt, escaping ejecta and surface organics. *Icarus* **175**, 522–533. (doi:10.1016/j.icarus.2004.12.005)

Asphaug, E., Agnor, C. B. & Williams, Q. 2006 Hit-and-run planetary collisions. *Nature* **439**, 155–160. (doi:10.1038/nature04311)

Benz, W., Slattery, W. L. & Cameron, A. G. W. 1986 The origin of the Moon and the single impact hypothesis I. *Icarus* **66**, 515–535. (doi:10.1016/0019-1035(86)90088-6)

Bowring, S. A. & Williams, I. S. 1999 Priscoan (4.00–4.03 Ga) orthogneisses from northwestern Canada. *Contrib. Mineral. Petrol.* **134**, 3–16. (doi:10.1007/s004100050465)

Cameron, A. G. W. 2002 Birth of a solar system. *Nature* **418**, 924–925. (doi:10.1038/418924a)

Canup, R. M. 2004 Simulation of a late lunar-forming impact. *Icarus* **168**, 433–456. (doi:10.1016/j.icarus.2003.09.028)

Canup, R. M. & Pierazzo, B. 2006 Retention of water during planet-scale collisions. *LPSC* **XXXVII**, abst. 2146.

Cassen, P. 2001 Nebular thermal evolution and the properties of primitive planetary materials. *Meteorit. Planet. Sci.* **36**, 671–700.

Chambers, J. H. 2001 Making more terrestrial planets. *Icarus* **152**, 205–224. (doi:10.1006/icar.2001.6639)

Chambers, J. H. 2003 Symplectic integrators with complex time steps. *Astron. J.* **126**, 1119–1126. (doi:10.1086/376844)

Chyba, C. F. 2000 Energy for microbial life on Europa. *Nature* **403**, 381–382. (doi:10.1038/35000281)

Chyba, C. F., Thomas, P. J., Brookshar, L. & Sagan, C. 1990 Cometary delivery of organic molecules to the early Earth. *Science* **249**, 366–373.

Cockell, C. 2006 The origin and emergence of life under impact bombardment. *Phil. Trans. R. Soc. B* **361**. (doi:10.1098/rstb.2006.1908)

Cyr, K. E., Sears, W. D. & Lunine, J. I. 1998 Distribution and evolution of water ice in the solar nebula: implications for solar system body formation. *Icarus* **135**, 537–548. (doi:10.1006/icar.1998.5959)

Donahue, T. M. 1989 Deuterium on Venus. *Nature* **340**, 513–514. (doi:10.1038/340513b0)

Drake, M. J. & Righter, K. 2002 Determining the composition of the Earth. *Nature* **416**, 39–44. (doi:10.1038/416039a)

Elachi, C. *et al.* 2005 First views of the surface of Titan from the Cassini RADAR. *Science* **308**, 970–974. (doi:10.1126/science.1109919)

Fajardo-Cavazos, P., Link, L., Melosh, H. J. & Nicholson, W. L. 2005 *Bacillus subtilis* spores on artificial meteorites survive hypervelocity atmospheric entry: implications for lithopanspermia. *Astrobiology* **5**, 726–736. (doi:10.1089/ast.2005.5.726)

Ganguly, J. & Bose, K. 1995 Kinetics of formation of hydrous phyllosilicates in the solar nebula. *LPSC* **XXVI**, 441–442.

Genda, H. & Abe, Y. 2005 Effects of oceans on atmospheric loss during the stage of giant impacts. *LPSC* **XXXVI**, abs. 2265.

Goldreich, P., Lithwick, Y. & Sari, R. 2004 Final stages of planet formation. *Astrophys. J.* **614**, 497–507. (doi:10.1086/423612)

Halliday, A. N. 2004*a* In *The origin and earliest history of the Earth* (ed. A. M. Davis, H. D. Holland & K. K. Turekian) Treatise on geochemistry, vol. 1, pp. 509–557. Amsterdam, The Netherlands: Elsevier Pergamon.

Halliday, A. N. 2004*b* Mixing, volatile loss and compositional change during impact-driven accretion of the Earth. *Nature* **427**, 505–509. (doi:10.1038/nature02275)

Hartmann, W. K. & Davis, D. R. 1975 Satellite-sized planetesimals and lunar origin. *Icarus* **24**, 504–515. (doi:10.1016/0019-1035(75)90070-6)

Kargel, J. F. 2006 Eceladus: cosmic gymnast, volatile miniworld. *Nature* **311**, 1389–1391.

Kasting, J. F. 1988 Runaway and moist greenhouse atmospheres and the evolution of Earth and Venus. *Icarus* **74**, 472–494. (doi:10.1016/0019-1035(88)90116-9)

Kasting, J. F. 2006 Atmospheric composition and climate on the early Earth. *Phil. Trans. R. Soc. B* **361**, 1733–1742. (doi:10.1098/rstb.2006.1902)

Knauth, L. P. 1992 Origin and diagenesis of cherts: an isotopic perspective. In *Isotopic signatures and sedimentary records* (ed. N. Clauer & S. Chandhuri), pp. 123–152. Berlin, Germany: Springer.

Levison, H. F. & Agnor, C. 2003 The role of giant planets in terrestrial planet formation. *Astron. J.* **125**, 2692–2713. (doi:10.1086/374625)

Lorenz, R. D., Lunine, J. I. & McKay, C. P. 2001 Geologic settings for aqueous organic synthesis on titan revisited. *Enantiomer* **6**, 83–96.

Lunine, J. I. 2005 *Astrobiology: a multidisciplinary approach*. San Francisco, CA: Addison-Wesley.

Lunine, J. I., Chambers, J., Morbidelli, A. & Leshin, L. A. 2003 The origin of water on Mars. *Icarus* **165**, 1–8. (doi:10.1016/S0019-1035(03)00172-6)

McCord, T. & Sotin, C. 2005 Ceres: evolution and current state. *J. Geophys. Res.* **110**, E05009. (doi:10.1029/2004JE002244)

Mojzsis, S. J., Harrison, T. M. & Pidgeon, R. T. 2001 Oxygen isotope evidence from ancient zircons for liquid water at the Earth's surface 4300 Myr ago Jack Hills, evidence for more very old detrital zircons in Western Australia. *Nature* **409**, 178–181. (doi:10.1038/35051557)

Moorbath, S. 2005 Oldest rocks, earliest life, heaviest impacts, and the Hadean–Archean transition. *Appl. Geochem.* **20**, 819–824. (doi:10.1016/j.apgeochem.2005.01.001)

Morbidelli, A., Chambers, J., Lunine, J. I., Petit, J. M., Robert, F., Valsecchi, G. B. & Cyr, K. E. 2000 Source regions and timescales for the delivery of water on Earth. *Meteorit. Planet. Sci.* **35**, 1309–1320.

Najita, J. & Williams, J. P. 2005 An 850 μm survey for dust around Solar–Mass stars. *Astrophys. J.* **635**, 625–635. (doi:10.1086/497159)

Niemann, H. B. *et al.* 2005 The composition of Titan's atmosphere from the GCMS on the Huygens probe, and implications for the origin of nitrogen and methane. *Nature* **438**, 779–784. (doi:10.1038/nature04122)

O'Brien, D. P., Lorenz, R. D. & Lunine, J. I. 2005 Numerical calculations of the longevity of impact oases on Titan. *Icarus* **173**, 243–253. (doi:10.1016/j.icarus.2004.08.001)

O'Brien, D. P., Morbidelli, A. & Levison, H. F. In press. Terrestrial planet formation with strong dynamical friction. *Icarus*.

Pascucci, I. *et al.* In press. Formation and evolution of planetary systems: upper limit on the gas mass in disks around Sun-like stars. *Astrophys. J.*

Pavlov, A. A., Kasting, J., Brown, L. L., Rages, K. A. & Freedman, R. 1990 Greenhouse warming by CH_4 in the atmosphere of early. *Earth J. Geophys. Res.* **105**, 11 981–11 981. (doi:10.1029/1999JE001134)

Peck, W. H., Valley, J. W., Wilde, S. A. & Graham, C. M. 2001 Oxygen isotope ratios and rare earth elements in 3.3. to 4.4 Ga zircons: ion microprobe evidence for high $\partial\,^{18}O$ continental crust and oceans in the Early Archean. *Geochim. Cosmochim. Acta* **65**, 4215–4229. (doi:10.1016/S0016-7037(01)00711-6)

Phillips, R. J. & Hansen, V. L. 1994 Tectonic and magmatic evolution of Venus. *Ann. Rev. Earth Planet. Sci.* **22**, 597–656. (doi:10.1146/annurev.ea.22.050194.003121)

Raymond, S., Quinn, T. & Lunine, J. I. 2006 High resolution simulations of the final assembly of Earth-like planets I: terrestrial accretion and dynamics. *Icarus* **183**, 265–282.

Robert, F. 2001 The origin of water on Earth. *Science* **293**, 1056–1058. (doi:10.1126/science.1064051)

Sasselov, D. D. & Lecar, M. 2000 On the snow line in dusty protoplanetary disks. *Astrophys. J.* **528**, 995–998. (doi:10.1086/308209)

Sleep, N. H. & Hessler, A. M. 2006 Weathering of quartz as an Archean climatic indicator. *Earth Planet. Sci. Lett.* **241**, 594–602. (doi:10.1016/j.epsl.2005.11.020)

Sleep, N. H. & Zahnle, K. 2001 Carbon dioxide cycling and implications for climate on ancient Earth. *J. Geophys. Res.* **106**, 1373–1400. (doi:10.1029/2000JE001247)

Sleep, N. H., Zahnle, K. J., Kasting, J. F. & Morowitz, H. J. 1989 Annihilation of ecosystems by large asteroid impacts on the early Earth. *Nature* **342**, 139–142. (doi:10.1038/342139a0)

Stein, C. A. & Stein, S. 1994 Constraints on hydrothermal heat flux through the oceanic lithosphere from global heat flow. *J. Geophys. Res.* **99**, 3081–3095. (doi:10.1029/93JB02222)

Stimpfl, M., Lauretta, D. S. & Drake, M. J. 2004. Adsorption as a mechanism to deliver water to the Earth. *Meteorit. Soc.*, Abstract.

Tobie, G., Lunine, J. I. & Sotin, C. 2006 Episodic outgassing as the origin of atmospheric methane on Saturn's moon Titan. *Nature* **440**, 61–64.

Tomasko, M. *et al.* 2005 Rain, winds, and haze during the Huygens probe descent to Titan's surface. *Nature* **438**, 765–778. (doi:10.1038/nature04126)

Valley, J. W., Peck, W. H., King, E. M. & Wilde, S. A. 2002 A cool early Earth. *Geology* **30**, 351–354. (doi:10.1130/0091-7613(2002)030<0351:ACEE>2.0.CO;2)

Veeder, G. J., Matson, D. L., Johnson, T. V., Davies, A. G. & Blaney, D. L. 2004 The polar contribution to the heat flow of Io. *Icarus* **169**, 264–270. (doi:10.1016/j.icarus.2003.11.016)

Wahr, J. M., Zuber, M. T., Smith, D. E. & Lunine, J. J. In press. Tides on Europa. *Geophys. Res.* (*planets*).

Ward, W. R. 1973 Large scale variations in the obliquity of Mars. *Science* **181**, 260–262.

Ward, P. & Brownlee, D. 2000 *Rare Earth*. New York, NY: Copernicus Books.

Zolotov, M. Y. & Shock, E. L. 2003 Energy for biologic sulfate reduction in a hydrothermally formed ocean on Europa. *J. Geophysical Res.* **108**, 3–1. CiteID 5022. (doi:10.1029/2002JE001966)